THE LONG-TERM STABILITY OF ENGINEERED LANDFORMS OF THE RANGER URANIUM MINE, NORTHERN TERRITORY, AUSTRALIA: APPLICATION OF A CATCHMENT EVOLUTION MODEL

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ABSTRACT

There is a need to assess the long-term stability of engineered landforms associated with the rehabilitation of Ranger Uranium Mine, Northern Territory, Australia, as it is a requirement that mill tailings must be contained for periods in excess of 1000 years. The geomorphic model, SIBERIA, is calibrated on hydrologic and erosion data collected by a combination of monitoring and rainfall simulation experiments on the waste rock dumps of Ranger. Preliminary analysis of Ranger's preferred above-grade option suggests that erosion of the order of 7 to 8 m will occur on the structure in a period of 1000 years. This depth of erosion may be sufficient to compromise the integrity of containment. It is shown that SIBERIA has significant advantages over steady-state erosion models. Suggestions are made for the design that will enhance the stability of the structure and extend the structural life of the containment. © 1998 Reproduced with permission of the Commonwealth of Australia.

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KEY WORDS: hillslope model; uranium mine rehabilitation; mine waste; erosion rates; Australia; Alligator Rivers Region

INTRODUCTION

The rehabilitation of the Ranger Uranium Mine, Northern Territory, Australia, will involve shaping waste rock dumps, consisting of more than 100 million tonnes of waste rock and low grade ore, and containing the mill tailings. The mill tailings may be rehabilitated either within the existing above-ground 'turkey-nest' tailings dam (the above-grade option) or encapsulated in the mine pits (the below-grade option). The containment structures must be structurally stable for a minimum period of 1000 years (Commonwealth of Australia, 1987), and it is essential that the stability can be demonstrated. Ranger Uranium Mines Pty Ltd (RUM) prefers the above-grade uption (Unger *et al*., 1989) but the Environmental Requirements (Commonwealth of Australia, 1979) specify a below-grade option. Engineered landforms, which include covers over the mill trailings and batter slopes, will be constructed from the waste rock, a chlorite-rich schist which weathers rapidly to gravel and finer fractions (Milnes *et al*., 1986; Riley, 1994). Geomorphic processes of weathering and erosion will largely determine the long-term stability of the structures (Riley, 1995).

A geomorphic model that predicts the long-term changes in landforms and water and sediment discharge is needed to assess the long-term stability (Riley, 1995). The research programme for developing and testing this model has involved detailed studies of erosion and hydrologic processes on waste rock and natural sites, as well as assessment of risk of dispersal of potential contaminants (Riley and Waggitt, 1992; Riley, 1994; Rippon and Riley, 1996). Monitoring and simulation was used to collect hydrogeomorphic data for examining the critical processes and providing a data set for calibration of hydrogeomorphic models. Details of the experimental techniques are given in Riley and East (1990), Riley and Gardiner (1991, 1992) and Riley (1994).

This paper assesses the long-term erosional stability of engineered landforms at the Ranger Uranium Mine using the Willgoose Catchment Evolution Model (SIBERIA; Willgoose *et al*., 1989, 1991a,b,c,d,e). specifically, SIBERIA is calibrated using hydrogeomorphic data collected by monitoring and rainfall simulation on the waste rock dumps of Ranger Uranium Mine and is used to predict the evolution of the

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proposed engineered landforms over 1000 years. Implications for the stability of the ‘below- and above-grade’ options are discussed.

LONG-TERM LANDSCAPE EVOLUTION MODEL (SIBERIA)

SIBERIA is a physically based model of the evolution of landforms under the action of fluvial erosion, creep and mass movement. A crucial component of SIBERIA is that it explicitly incorporates the interaction between the hillslopes and the growing channel or gully network based on physically observable mechanisms. The elevations within the catchment – both hillslope and channel – are simulated by a mass-transport continuity equation applied over geologic time. Mass-transport processes considered include fluvial sediment transport, such as modelled by the Einstein–Brown equation, and mass-movement mechanisms such as creep, rainsplash or landslide. An explicit differentiation between the processes that act on the hillslopes and in the channels is made. The growth of the channel network is governed by a physically based threshold mechanism, where if a function (called the channel initiation function) is greater than some predetermined threshold then channel head advance occurs. The channel initiation threshold is dependent on the resistance of the catchment to channelization. Channel growth is thus governed by the hillslope form and processes that occur upstream of the channel head, but in a way that can be independent of channel growth stability arguments (Smith and Bretherton, 1972). The elevations on the hillslopes and the growing channel interact through the different transport processes in each regime and the preferred drainage to the channels that result. The interaction of these processes produces the long-term form of the catchment. The preferential erosion in the channels results in the familiar pattern of hills and valleys with hillslope flow being towards the channel network in the bottoms of the valleys.

The mass-transport processes in SIBERIA include both fluvial sediment transport and a conceptualization of diffusive mass movement mechanisms such as creep, rainsplash and landslide. The model averages these processes in time so that the elevations (and the channel network) are indicative of the average, with time, of the full range of erosion events; the elevations simulated are average elevations with time. The mathematical details of this equation are discussed elsewhere (Willgoose *et al*., 1991a, b, c, d).

SIBERIA is a flexible framework within which to study the evolution of landforms. The computer code is structured so that any user-defined erosion, channel, runoff, tectonic and geology model that is dependent on elevation either directly or indirectly (through, for example, contributing areas or slopes) can be incorporated. Parameters may vary in space and time as desired. At the present the time-stepping algorithm is optimized for fluvial sediment transport with a functional form as given in Equation 2 (Willgoose *et al*., 1991a). The following discription of SIBERIA concentrates solely on the capabilities used in this study. A more complete description of its capabilities is available in Willgoose (1992).

The evolution of the landform at a point follows directly from the mass conservation of sediment:

$$\frac{\partial z}{\partial t} = c_0 + \frac{\nabla q_s}{\rho_s (1-n)} + \nabla q_m \tag{1}$$

where z = elevation, t = time, c_0 = tectonic uplift, ∇ = vector differential, q_s = sediment transport per unit width (mass/time/unit/width), q_m = mass movement transport per unit width (mass/time/unit/width), ρ_s = density of the sediment and n = porosity of the sediment.

The first term in this equation is the rate of tectonic uplift (positive upwards). The third term represents diffusive mechanisms occurring in certain mass-transport processes, such as creep, rainsplash and landsliding (Culling, 1963; Dunne, 1980; Andrews and Bucknam, 1987). Both the diffusivity and tectonic uplift may vary in time and over the catchment. In principle, it is possible to use more sophisticated diffusive models of hillslope evolution (Ahnert, 1976) but at this time data do not appear to be available to define them accurately. These enhancements (e.g. viscous and plastic flow) are typically spatially variable and dependent on soil depth. At this time SIBERIA does not model the chemical and physical processes associated with weathering and soil formation. Accordingly, it cannot model those processes that depend on soil depth.

The sediment transport process, q_s, modelled by the second term in Equation 1, can be parameterized in any

way that is believed to reflect the processes occuring in the catchment. Willgoose *et al*., (1989) and others (e.g. Kirkby, 1971) suggest that a realistic formulation is of the form:

$$q_s = f(Y)Gq^{m_1}S^{n_1} \tag{2}$$

where q = discharge per unit width, S = slope in the steepest downstream direction, $f(Y)$ = a sediment transport coefficient dependent on the pattern of channelization (discussed below), G = a function dependent on the runoff process modelled (discussed below) and m_1, n_1 = sediment transport coefficients. This fluvial sediment transport term is one that has been commonly used by geomorphologists (Kirkby, 1971; Smith and Bretherton, 1972; Moore and Burch, 1986) to represent a transport-limited process. The equation is used to model the mean sediment transport, so that the appropriate discharge to use is the mean peak discharge derived from a frequency analysis of runoff events (Willgoose *et al*., 1989).

The function G indicates what proportion of storms saturate that point and thus generate surface runoff. It is only during those storms when surface runoff is generated that fluvial erosion occurs. The calculation of this parameter is discussed in further detail in Willgoose *et al*. (1991e). Suffice to say that for Hortonian runoff it might be assumed that G = 1. For subsurface saturation-generated runoff, smaller storms saturate a smaller proportion of the catchment than do larger storms. Thus for subsurface saturation runoff G is less than 1 and largest in those parts of the catchment saturated most frequently. In this paper only Hortonian runoff is considered so that G = 1.

The slope in the fluvial sediment transport equation is determined from the catchment elevations and direction of steepest downhill drainage. The discharge relationship, dependent on area and slope, can be formulated to reflect the processes that occur in the field. However, it is important to note that if the sediment transport equation is to model the long-term average sediment transport equation then the discharge per unit width q, should be interpreted as the mean annual peak discharge, analogous to the idea of a dominant discharge (Willgoose *et al*., 1989), so that:

$$Q = \beta_3 A^{m_3} S^{n_3} \tag{3}$$

where Q = discharge in the channel, β_3 = runoff rate constant, A = area per unit width and m_3, n_3 = coefficients. This empirical relationship accounts for runoff routing effects within the catchment and the spatial correlation of rainfall (Leopold *et al*., 1964; Pilgrim, 1987) and also the observed behaviour of kinematic wave-based rainfall-runoff models (Willgoose and Kuczera, 1995).

A feature of this model is its ability to explicitly model the dynamics of the channel network, extension and retreat, and to allow for different sediment transport processes in the channels and on the hillslopes. A variable is defined, Y, that identifies where channels exist (Y = 1) versus where the catchment is hillslope (Y = 0). Initially a catchment can have either no channels or a predefined channel and drainage network. The extension of the channel network occurs when a function, non-linearly dependent on contributing area per unit width and slope, called the channel initiation function a, exceeds a threshold value called the channel initiation threshold a_t. Various options are available that allow for channel retreat when the channel initiation function is small, or alternatively, a channel once formed can be permanent. The exact means by which the transformation between hillslope and channel occurs appears to be unimportant. More important is the functional dependence of the channel initiation function on discharge and slope which Willgoose and co-workers have formulated as:

$$a = \beta_5 q^{m_5} S^{n_5} \tag{4}$$

where a = channel initiation function and β_5, m_5, n_5 = coefficients.

Again, within the conceptual framework of the model, the form of the channel initiation function can be formulated as seen fit in light of physical processes observed in the field. This particular formulation can model

a range of physical processes (Willgoose *et al.*, 1989; Montgomery, 1994) and it is consistent with field data collected by other workers (Patton and Schumm, 1975; Montgomery and Dietrich, 1989).

The channel network calculated by the model is used to determine the rate at which fluvial sediment transport occurs as:

$$f(Y) = \begin{cases} \beta_1 \\ \beta_1 O_t \end{cases} \tag{5}$$

where β_1 = erosion rate constant and O_t = ratio of hillslope to channel erosion rate.

The transport rate β_1 can be spatially and temporally variable so that, for instance, structural controls due to the differential erodibility of strata can be easily modelled. Diffusive transport can also be spatially variable. Channel depth and width can be modelled with a function (regime equation) dependent on area and slope (Leopold *et al.*, 1964; Riley and Williams, 1991). Consistency of elevations and slopes between channels and hillslopes is maintained and sediment fluxes resulting from excavation of channels are accounted for in Equation 1.

HYDROLOGY MODEL

Runoff is the most important determinant of soil erosion. Thus accurate simulation of runoff is necessary to provide a reliable erosion assessment. The hydrology model used to fit to the rainfall simulator and natural rainfall plots is based on the 1-D kinematic wave flood routing model described by Field and Williams (1987), the Generalized Kinematic Catchment Model (GKCM). The model has been extended to allow it to use a digital terrain map as the basis for determining flow paths and velocities and the scaling properties of the simulated runoff examined (Willgoose and Kuczera, 1995). For this study there were two relevant modifications to the model.

1. The groundwater component on the model has been disabled. Infiltration is assumed to drain to a very deep aquifer which does not discharge to the surface within the study site. This is believed to be a good representation of the waste rock dump (Woods, 1994).
2. The area associated with each subcatchment (node) is the same, equal to the square of the grid spacing. The digital terrain models (DTMs) provided for the waste rock dump are on a 30 m grid so that the area of each cell is 900 m^2.

EROSION MODEL

The overland flow erosion model used to fit the erosion data is:

$$q_s = \beta_1' q^{m_1'} S^{n_1'} (\tau - \tau_c) \tag{6}$$

where τ = bottom shear stress for the flow, and τ_c = shear stress threshold. The parameters β_1', m_1' and n_1' are fixed by the flow cross-section geometry and erosion physics. This equation parameterizes the total load, i.e. the sum of both the suspended and bed loads. For instance, for a constant-width channel with no downstream trends in sediment characteristics, and with sediment transport according to the Einstein–Brown equation, $m_1'=1{\cdot}8$ and $n_1'=2{\cdot}1$. For flow in rills the parameters are approximately $m_1'=1{\cdot}3$ and $n_1'=2{\cdot}2$ (Moore and Burch, 1986; Willgoose *et al.*, 1989). Exact values for these parameters depend on the rill cross-section. The parameter β_1' gives the rate of sediment transport and is primarily a function of sediment grain size, vegetation cover and land use, analagous to the *K* factor in other erosion models (e.g. CREAMS, USLE).

Riley (1992) attempted to identify a shear stress threshold as in Equation 6 for the material from the waste rock dump using a small flume and concluded that the value was very small, and that he was unable to reliably estimate it. On this basis the shear stress threshold in all the work that follows will be assumed to be zero. We

also note that bottom shear stress, τ, can be described by a function of discharge and slope (Willgoose *et al*., 1989) so that the sediment transport model that is used here has the form:

$$q_s = \beta_1 q^{m_1} S^{n_1} \tag{7}$$

Similarly to Equation 2, this model does not incorporate hysteretic effects in the sediment rating curve that may result from sediment storage. However, since there are no data for the region that indicates the possible importance of this effect over areas of the size of the waste rock dump, hysteresis has been ignored. It is not possible at present to predict the magnitude of this effect without extensive field data for large catchments.

For smaller areas the overland erosion is dominated by rainsplash, or rain-flow, erosion. Rainsplash is generally modelled by an additive Fickian diffusion term where the diffusivity, D, is a function of the applied rainfall energy (in turn, a function of the energy of the individual raindrops and the rainfall rate) so that:

$$D = \alpha R$$

where R = the rainfall rate.

The mean concentration in the flow, c, for the area above the point of interest, is then given by:

$$c = \beta_1 q^{m_1 - 1} S^{n_1} + \frac{DS}{q} = \beta_1 q^{m_1 - 1} S^{n_1} + \frac{D' RS}{q} \tag{8}$$

These equations indicate that as the discharge decreases, or the area of the plot decreases, then the second, diffusive term will begin to dominate the expression for concentration. For large areas the diffusive term becomes relatively small so that Equation 2 governs. That the diffusive term is additive implies that the processes that cause diffusive transport and those that cause fluvial transport do not interact. Thus a higher or lower level of diffusive transport does not of itself change the rate of fluvial transport.

Willgoose *et al*. (1989) showed that if the sediment transport rate was described by Equation 7 then the long-term average could also be expressed in that form though not, in general, with the same exponents. The interpretation of β_1 and q are modified (β_1 is a frequency factor and q is the mean peak discharge) and the runoff modelling of this report is aimed at simulating these runoff data for the waste rock dump.

The model of Equation 6 is primarily used for ‘transport-limited’ sediment transport. That is, it is assumed that there is always sufficient material on the surface to satisfy the transport demands of the flow. This is not the case when sediment starvation or source limitation occurs and it is not established that Equation 6 is necessarily complete for this case. There appears to be some evidence to suggest that sediment starvation can occur at Ranger. The issue of armouring and sediment starvation are matters of additional research. Riley (1992) showed that for a constant discharge the concentration of sediment decreased over a period of 1 h but the effect appeared to be small. Moreover, some of the concentration data for the natural rainfall events exhibited clockwise rating curves with discharge. However, if the sediment transport model is calibrated to the natural data, rather than simulated runoff data, this effect will be accounted for, on average, in the calibration. For natural runoff events the recorded data will be for the naturally sediment-starved flow so that the calibrated β_1 will be lower, reflecting the sediment starvation. If, however, the sediment transport model is calibrated with the rainfall simulator trials then sediment starvation has to be accounted for explicitly. This study uses the natural data wherever possible to circumvent this problem. In any event, the small differences between the simulated rainfall and natural rainfall concentration data appear to have negligible effect on the relationship between discharge and concentration.

HYDROLOGY MODEL CALIBRATION

Monitoring and simulation experiments were conducted on two surfaces of Ranger Uranium Mine, the cap and the batter surfaces (Figure 1). The monitoring experiments involved two major instrumented areas, one on the cap and the other on the batter slope. At each site a number of wash and rill traps were installed. On the cap site it

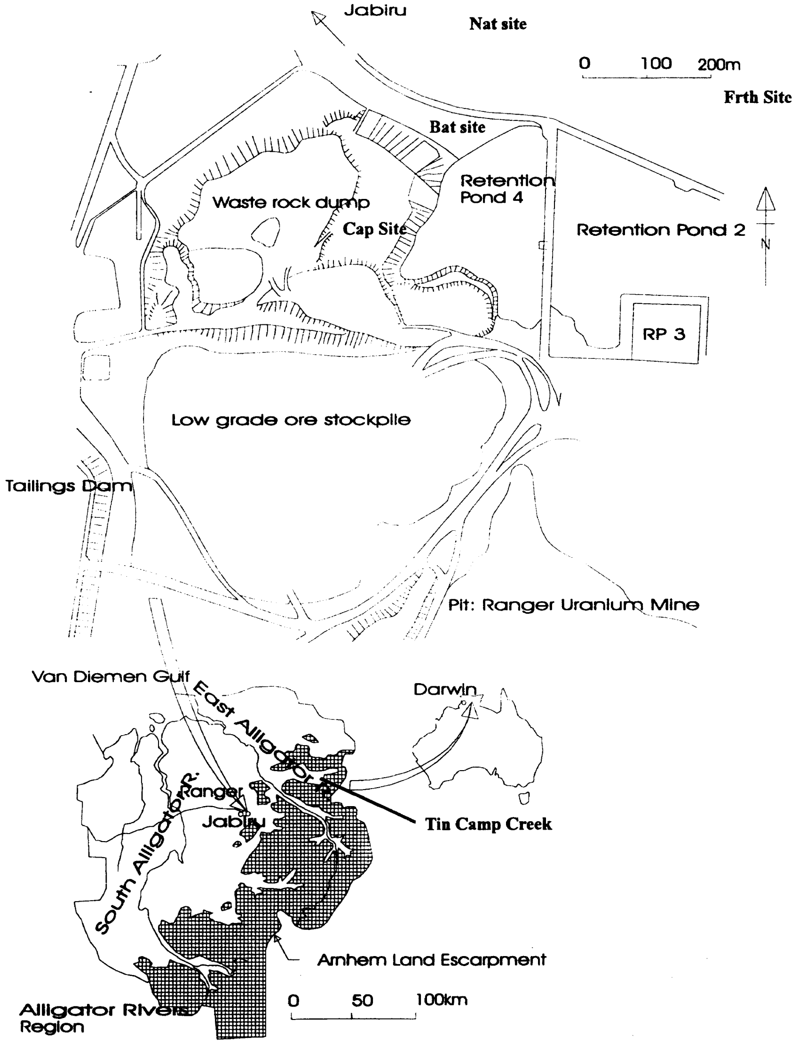


Figure 1. Location of the cap and batter sites at Ranger Uranium Mine

was possible to identify a major outlet and this was monitored for discharge and sediment load. Tipping bucket raingauges, manual raingauges, piezometers, tensiometers, erosion pins and painted stone lines were installed as part of the monitoring. Rainfall simulation used the simulator designed by Riley *et al.* (in prep). Multiple plots were monitored during the simulation runs, ranging in size from $1\,m^1$ to $>100\,m^2$. Further details of the monitoring and simulation experiments are given in Riley and Gardiner (1991, 1992) and Riley (1994).

The cap is a low gradient surface (1 to 5 per cent) which constitutes the majority of the surface of the waste rock. The batter slope surface is composed of the steep batter slopes (10 to 30 per cent) that surround the waste rock dump. In both cases the surfaces were not rehabilitated and had not been disturbed since being laid down during the dumping. Previous studies have shown that the surfaces are compact, have low infiltrabilities and are covered with a layer of fine to medium platty gravels (Williams *et al.*, 1990; Gardiner *et al.*, 1990; Riley and Gardiner, 1991; Riley, 1994). Table I summarizes the runoff data that have been used in this study. Further details on the data used in the calibration are provided in Willgoose and Riley (1993a). The plot characteristics are listed in Table II.

The primary data used for calibration of the rainfall–runoff model were the natural rainfall events. Reliable events were selected for several sites and the model parameters adjusted by trial and error to give a good fit. The

Table I. Monitoring data supplied for caprock and batter sites

Date of storm	Cap sites					Batter sites		
	WT1	WT2	WT3	RT1	OUT	RT2	WT1	WT2
7/1/91 (20:50)*					2†			
7/1/91 (14:55)*		2		2	2†			
8/1/91					2†			
10/1/91 (7:55)*	2	2		2	2†			
10/1/91 (14:00)*					2†			
11/1/91					2†			
21/1/91	2	2						
28/1/92					2†			
30/1/91								2
4/2/91		2					2	
6/2/91			2			2	2	2
13/2/91						2		
16/2/91	2		2			2	2	2
22/2/91						2	2	

Site notation: WT = wash trap; RT = rill trap; OUT = main catchment outlet
Notation: 2 = data appear to be accurate
* Two events for these dates; approximate start time in parentheses
† Data truncated above discharge $15\,l\,s^{-1}$

Table II. Cap site catchment characteristics

	Area (m^2)	Mean slope	Mean width (m)	Length (m)
COUT	2182	0·03	*	*
CRT1	461	0·029	*	*
CRT2	330	0·039	*	*
CRT3	731	0·034	*	*
CWT1	149*	0·04	4·57	32·6
CWT2	102	0·035	1·87	54·4
CWT3	91	0·036	1·63	55·6

* Variable width and length

broad range of hydrographs (single-peaked versus double-peaked hydrographs for a variety of closely spaced sites) exercised all components of the model. The rainfall simulator trials had less variation in discharge and did not exercise all the components of the model, making it difficult to reliably estimate their parameters. They provide useful verification data.

As far as possible, parameters were determined from, or checked against, other independent data. For instance, Manning *n* values for the kinematic wave routing have been checked against measures of surface roughness (Henderson, 1966) with a satisfactory correspondence.

Where extra runoff event data were available, verification of the selected model parameters was carried out. This verification was an important part of the process of ensuring that the selected model and parameters were satisfactory. The goodness of fit for the verification sites and events was, as expected, not quite as good as for the calibration runs, but showed a good correspondence in volume, peak discharge, time to peak, and an overall shape of the hydrograph. Details of the calibration are given in Willgoose and Riley (1993a).

Cap sites calibration: natural rainfall data

The first of the plots to be calibrated was CWT2. Three reliable storms were chosen and parameters fitted by trial and error. The only difference in the fitted parameters for the three storms were initial soil wetness conditions, parameterized by the initial sorptivity. The parameters adopted are given in Table III. The simulations of runoff for the three storms are given in Figures 2, 3 and 4. The chosen parameters fit all aspects of the hydrographs well. In all cases the timing of peaks is satisfactory, indicating that the conveyance parameters (e.g. Manning *n*) are satisfactory. The widths of the hydrographs are satisfactory indicating that the surface storage parameters are satisfactory. Finally, the volume and the distribution of this volume within the hydrographs are well matched indicating that the infiltration parameters are satisfactory. Several other sites were also calibrated yielding similar parameters. Site CWT1 was examined independently to validate the model

Table III. Adopted parameters for the distributed Field–Williams model

Parameter	Value	(Range)
Sorptivity, initial infiltration, S_ϕ	$3{\cdot}85\,\mathrm{mm\,h^{-1}}$	(0–3·85)
Long-term infiltration rate, ϕ	$6{\cdot}5\,\mathrm{mm\,h^{-1}}$	(6·5–40)
Manning coefficient, n	0·03	(0·025–0·035)
Conveyance exponent, e_m	1·66	
Surface storage coefficient, c_r	0·03	
Surface storage exponent, γ	0·375	

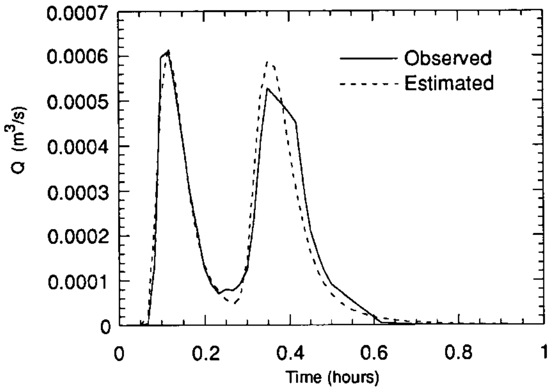


Figure 2. Calibration for CWT2 on 7/1/91

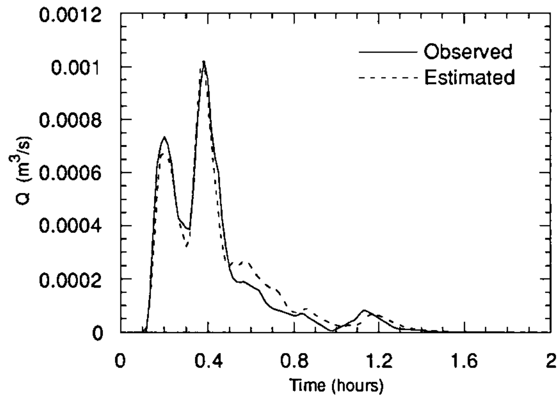


Figure 3. Calibration for CWT2 on 10/1/91

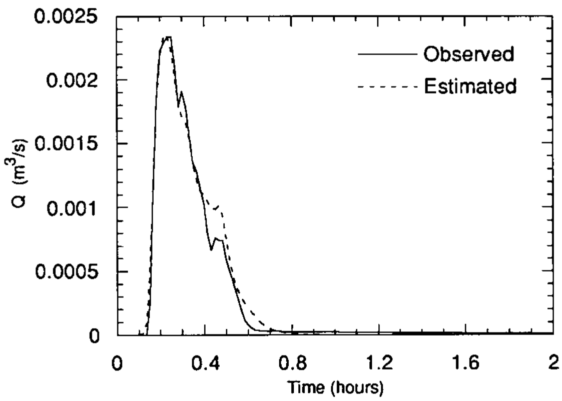


Figure 4. Calibration for CWT2 on 21/1/91

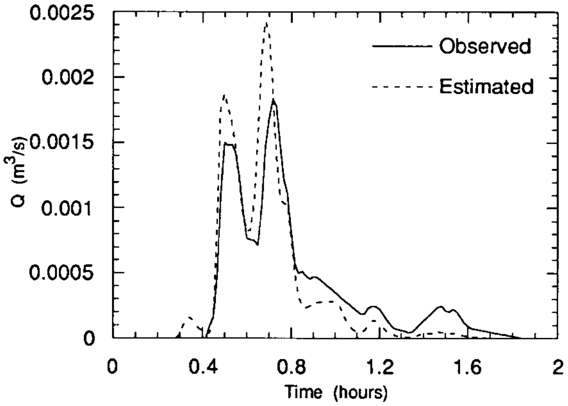


Figure 5. Verification for CWT1 on 10/1/91

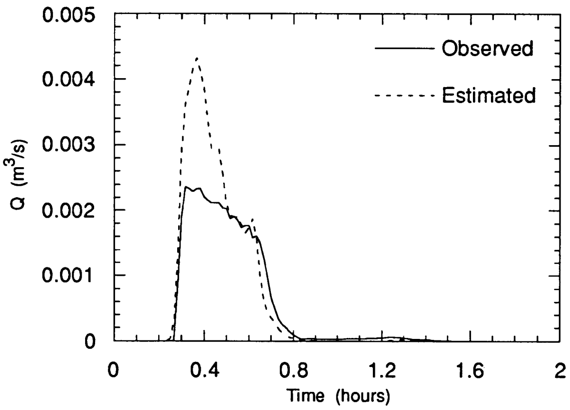


Figure 6. Verification for CWT1 on 21/1/91

parameters. This site had a number of rainfall events that were common with the calibration events described above. To validate the model it was decided to use the parameters and the initial conditions fitted for CWT2 to generate the events on CWT1. No parameters are fitted. The two events used were 10/1/91 and 21/1/91. The observed and predicted runoffs are given in Figures 5 and 6. An initial peak in the event of 21/1/91 appears in the simulation that is not evident in the data, otherwise the fit of the simulation data to the runoff data for these two events appears to be satisfactory, providing confidence in the validity of the parameters.

Conclusion

The distributed Field–Williams rainfall–runoff model (DISTFW) has been shown to provide a satisfactory fit to the data collected on the caprock and batter sites. Scatter in the infiltration parameters was observed from site to site, possibly suggesting localized porous zones. The sorptivity also exhibited small fluctuations depending on whether a rainfall event had occurred previously that day. This variation in the sorptivity appears to be important but its variation with antecedent wetness conditions could only be estimated roughly in this study because of the lack of detailed data testing specifically for antecedent wetness. The adopted parameters are listed in Table III.

EROSION MODEL CALIBRATION

The instantaneous sediment transport model described by Equation 7 was calibrated against the sedimentological data. Tables IV and V summarize the sediment yield data used. There are four parameters that require determination: β_1, m_1, n_1 and D, the transport rate, the exponents on the discharge and slope respectively,

Table IV. Sediment yield data supplied for caprock and batter sites – monitoring [(c)]

Date of storm	Cap sites						Batter sites		
	CWT1	CWT2	CWT3	CRT1	CRT2	COUT	BRT2	BWT1	BWT2
7/1/91 (20:50)†	*								
7/1/91 (14:55)†		*		*					
21/1/91	*	*							
4/2/91		*						*	
6/2/91			*				*	*	*
13/2/91							*	*	
16/2/91	*		*				*	*	*
22/2/91							*	*	

Site notation: WT = wash trap; RT = rill trap; OUT = main catchment outlet
Notation: * = data appeared to have reliable matching discharge data
† Two events on this date; approximate start time in parentheses

Table V. Rainfall simulation data

Run	Date	Area (m^2)	Slope
plot 1 (runs 1–5), uncovered	5–7/10/90	1·2	0·152
plot 2 (runs 1–5), covered	5–7/10/90	0·99	0·185
plot 1 (runs 1–3)	17–18/10/91	1·2	0·152
plot 2 (runs 1–3)	17–18/10/91	10·2	0·187
plot 3 (run 1)	17/10/91	0·99	0·99
plot 4 (runs 2–3)	18/10/91	107·7	0·19
plot 1 (runs 2–4)	23–25/4/90	113·5	0·021
plot 2 (runs 2–4)	23–25/9/90	116·5	0·025
plot 1 (runs 1–5)	22–25	7·8	
plot 2 (runs 1–5)	22–25	4·5	

and the diffusivity of rainsplash, respectively. Multiple regression was used on the available data, over a range of discharges, catchment areas and slopes, to estimate these parameters.

Natural rainfall data

The result of the multiple regression analysis with the data that were considered reliable yielded the relationship:

$$c = 0 \cdot 27 q^{0 \cdot 22} S^{0 \cdot 01}; \quad r^2 = 0 \cdot 29 \tag{9}$$

The correlation coefficient indicates that the overall fit of this expression is quite poor. Figures 7 and 8 show the concentration data against the discharge and slope. Partial regression tests indicate that the exponent on discharge is significantly different at the 5 per cent level from 0, as is the multiplicative constant, while that for slope is not. Finally, there was no significant difference between the sediment concentrations for the wash traps and the rill traps.

A common characteristic of the data was that if the rising limb at the start of the storm had been observed, the first datum point at the start had an anomalously high concentration. This behaviour is not uncommon and is commonly believed to result from rainfall detachment of particles at the start of the event, while depressions are being filled and before runoff has begun. This hyperconcentrated water is then flushed in the first minutes of the storm after which concentrations fall back to normal levels (Loch, pers. comm.). The total mass of sediment in that first flush was not a significant proportion of the total mass of the event, and because these data bias the multiple regression procedure they were ignored in the fitting of Equation 9.

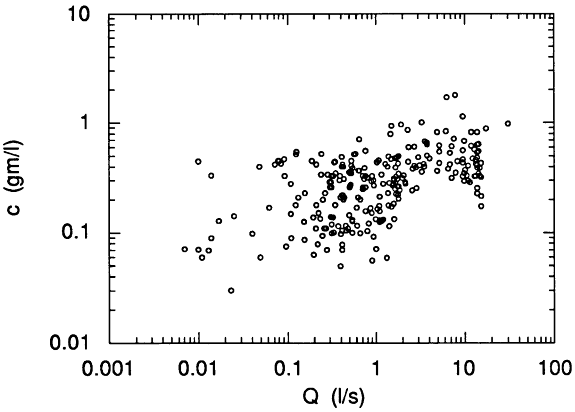


Figure 7. Concentration versus discharge for the natural rainfall events on both the batter and the cap rock sites

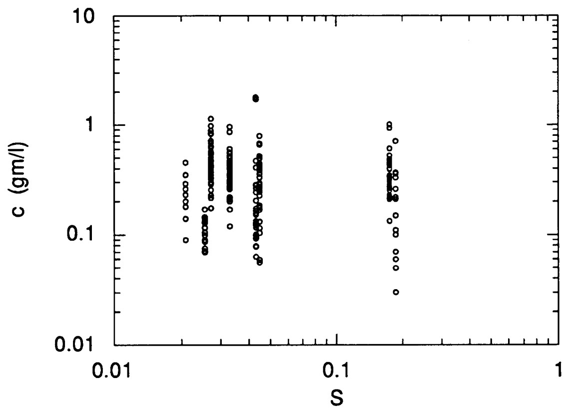


Figure 8. Concentration versus slope for the natural rainfall events on both the batter and the cap rock sites

Simulator rainfall data

The larger simulator catchments from the Batter and Cap plot 4 were used to calibrate the fluvial sediment transport equation of Equation 7 on the presumption that they would be least dominated by rainsplash. In that way multiple regression could be used to fit directly the parameters on discharge and slope. Figures 9 and 10 show the variation of the concentration with the discharge and slope for these large plots. The result of the multiple regression was:

$$c = 3 \cdot 55 Q^{0 \cdot 42} S^{0 \cdot 66}; \quad r^2 = 0 \cdot 62 \tag{10}$$

The correlation coefficient indicates that the fit of this equation is better than that obtained with the natural rainfall data. The exponent on discharge is consistent with fluvial transport according to Einstein–Brown sediment transport on a rilled surface (about 0·3–0·5; Willgoose *et al.*, 1989) and other field data (Loughran, 1977; Moore and Burch, 1986). The variation of concentration with discharge for one rainfall simulation experiment is illustrated on Figure 11. Note that the apparent decline in concentration with time is strongly associated with the period of initially increasing discharge and that this apparent starvation effect appears to be complete within about 20 min. Thereafter, concentration decreases with decreasing discharge as expected in the non-starved case.

The exponent on the slope in Equation 10 is considerably less than expected for a rilled surface (about 1·5–2). One possible explanation for this deviation is that the batter (hence higher slope) plots may have had a coarser lag layer (compared with the cap rock plots) so reducing the transport rate on the higher slope surfaces. The

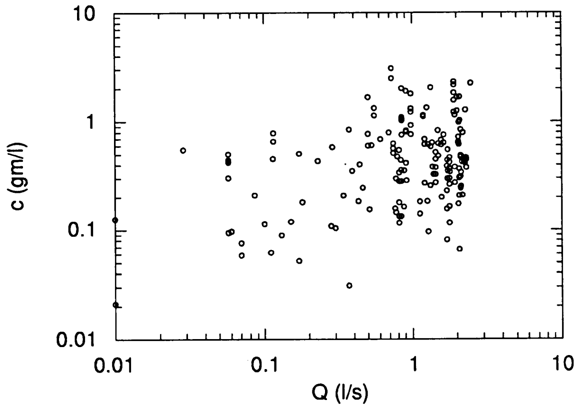


Figure 9. Concentration versus discharge for the large plots

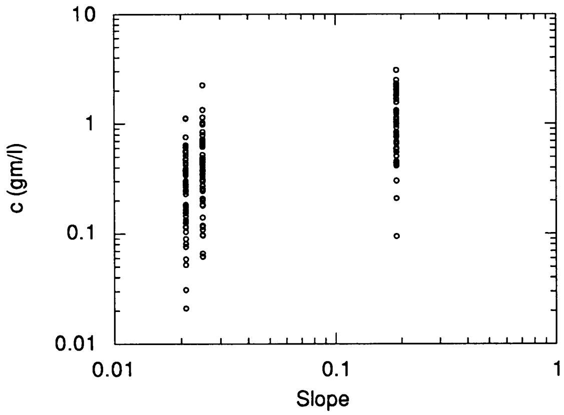


Figure 10. Concentration versus slope for the large plots

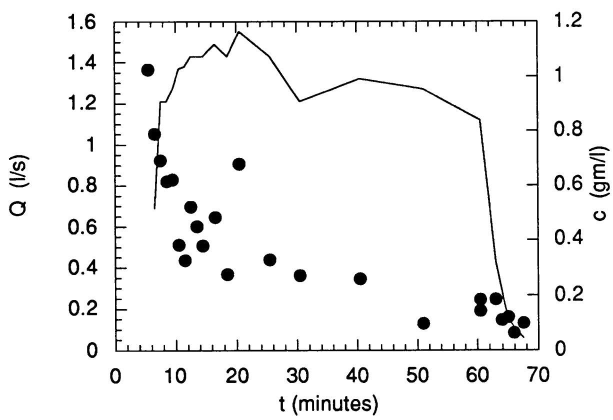


Figure 11. Sample hydrograph and sediment samples for cap rock; large plot rainfall simulation

expected exponent of 1·5–2 is derived using the assumption that the material grading properties do not change with discharge, slope or catchment area. Grading of samples taken from the cap rock and batter regions suggest that there could be minor differences in the lag layer, but the results were inconclusive.

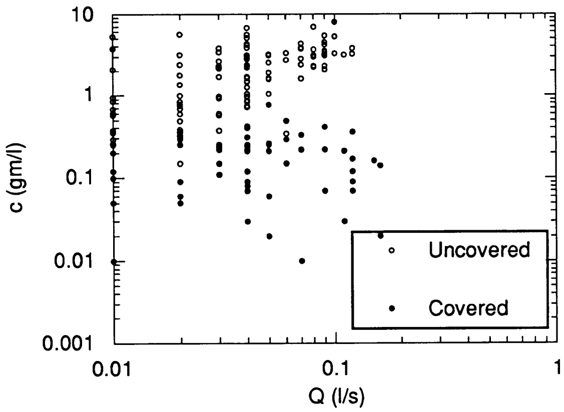


Figure 12. Concentration versus discharge for the covered and uncovered plots

Rainsplash diffusivity estimation

Having first calibrated the fluvial erosion on the larger catchments, the diffusive transport was calibrated to the data, which were fitted using multiple regression with trial and error estimates for the diffusive transport by using the transformation:

$$c-\frac{DS}{q}=\beta_1 q^{m_1-1}S^{n_1} \qquad (11)$$

The best estimate that was obtained by this process was:

$$c=3\cdot 59q^{0\cdot 68}S^{0\cdot 69}+\frac{0\cdot 178RS}{q};\quad r^2=0\cdot 638 \qquad (12)$$

The estimated parameters are only slightly different from those determined from the case ignoring diffusive transport in Equation 10, with a small improvement in the correlation coefficient for this case. The exponent on the discharge was significantly increased, though the value is still in the expected range for rill flow. The exponent on slope was not significantly changed.

The diffusivity obtained for the small plot data for the batter (plots 1 and 2, Table V) were used to independently check the diffusivity calculated above. These plots have comparable area and slope and vary only in that one was covered with netting and one was not. The purpose of the covering was to break the fall of the raindrops and thus dissipate the rainsplash energy. On small plots ($1\,m^2$) rainsplash is expected to dominate overland erosion. The difference in the transport of the two plots could thus be expected to be a good indicator of the magnitude of the rainsplash transport and thus the diffusivity. The concentration versus discharge for the two plots is illustrated in Figure 12. This change in concentration for the two plots can be expressed as:

$$\Delta c=\beta_1 q^{m_1-1}S^{n_1}-\left(\beta_1 q^{m_1-1}S^{n_1}+\frac{DS}{q}\right)=\frac{DS}{q} \qquad (13)$$

The calculation gave a value of $D=0\cdot 26$. This value is in good agreement with the value calculated from the large area plots of $D=0\cdot 178$. It is worthwhile to note that the plots' behaviour was consistent with the Fickian diffusion mechanism with transport being independent of discharge, validating its use for modelling small-scale sediment transport behaviour.

Finally, the data from a set of rainfall simulator plots mid-range in area between the small plots ($1\,m^1$) and large plots ($100\,m^2$), were examined as they were expected to exhibit behaviour midway between that of the

small and large plots. These mid-size plots, the ‘meso’ plots (5–8 m^2) exhibited this behaviour. For the lower rainfall rates (and thus lower discharges) rainsplash (diffusive) transport was dominant, with transport independent of discharge. For higher rainfall rates the behaviour appeared more like the fluvial transport mechanism, with transport discharge-dependent. However, it was noted that, if all the meso data were plotted together, there was a clear downward trend in concentration with increasing discharge. This appears to be due to sediment starvation, as the meso experiments were performed over four days, with increasing rainfall rates (yielding large discharges) being applied each day. It is believed that each day’s erosion was starved because of the depletion of the sediment store that had occurred with the previous day’s experiments. It is important, however, to observe that during each high rainfall rate experiment there was a positive correlation with discharge, consistent with all the results for the other plots discussed above. It is asserted here that the parameters m_1, n_1 and D fitted above are adequate for describing sediment transport during any event, but that the parameter β_1 may vary from event to event reflecting the amount of sediment removed from storage by runoff events of previous days.

Conclusion

The natural rainfall data and the simulator rainfall data have a significantly different functional dependence on slope, though a similar relationship with discharge. The problem remains as to which are the most reliable data. To this end the two sets of data were aggregated and examined as a whole. The concentrations for the natural rainfall experiments are lower than the simulator results, particularly for high slopes. This behaviour is marked and suggests a bias in the natural data, possibly due to sediment starvation, as previously discussed. In the graph of concentration versus discharge the results for the natural rainfall and simulated rainfall are little different, with the low discharge values for the natural event appearing to plot only slightly higher than the simulator.

DETERMINATION OF PARAMETERS FOR SIBERIA

The parameter estimation of the hydrology and sediment transport models described in the previous section provide the basis for estimation of the parameters for SIBERIA. The parameters of SIBERIA characterize the time-averaged properties of the processes, not the instantaneous or point values as calibrated in the hydrology and erosion studies above. However, there is very good reason to believe (Willgoose *et al.*, 1989; Huang and Willgoose, 1993; Willgoose and Kuczera, 1995) that the requisite average parameters can be obtained from the hydrology and erosion models calibrated in previous sections. The parameters in SIBERIA can be considered in two groups.

The first group of parameters in SIBERIA defines how the erosion varies with time, over periods of many years. This involves the averaging of the erosion that occurs in each runoff event to give the mean annual sediment yield. This mean annual sediment yield is not simply dependent on the sediment transport rate for a particular discharge and slope but also the range of discharges occurring during individual runoff events and the frequency at which these runoff events occur. Willgoose *et al*. (1989) have shown that the simple concentration–discharge–slope dependence calibrated above is maintained in the mean annual formula but that the discharge used in the equation changes from being the discharge at that time to the mean peak discharge obtained from a frequency analysis of runoff events. This peak discharge can in turn be related to the contributing area to that point. This mean peak discharge is very similar in interpretation to the dominant discharge, or channel-forming discharge, commonly used by river engineers in river sediment transport studies. The process that is followed is to simulate, using the hydrology model and observed pluviograph records for Jabiru, a runoff and erosion time series. The resulting erosion series is averaged over the simulated record and the average sediment transport rate is related to the mean peak discharge estimated by the hydrology model. This is similar to the recommended procedure for calibrating the erosion model CREAMS, though it normally uses daily data for sediment transport aggregation.

The second group of parameters defines how the hydrology changes at different points within the catchment and, in particular, how the mean peak discharge varies with area: the scale dependence of the runoff hydrology. The hydrology model uses the digital terrain map of the proposed mine sites to simulate the variation of

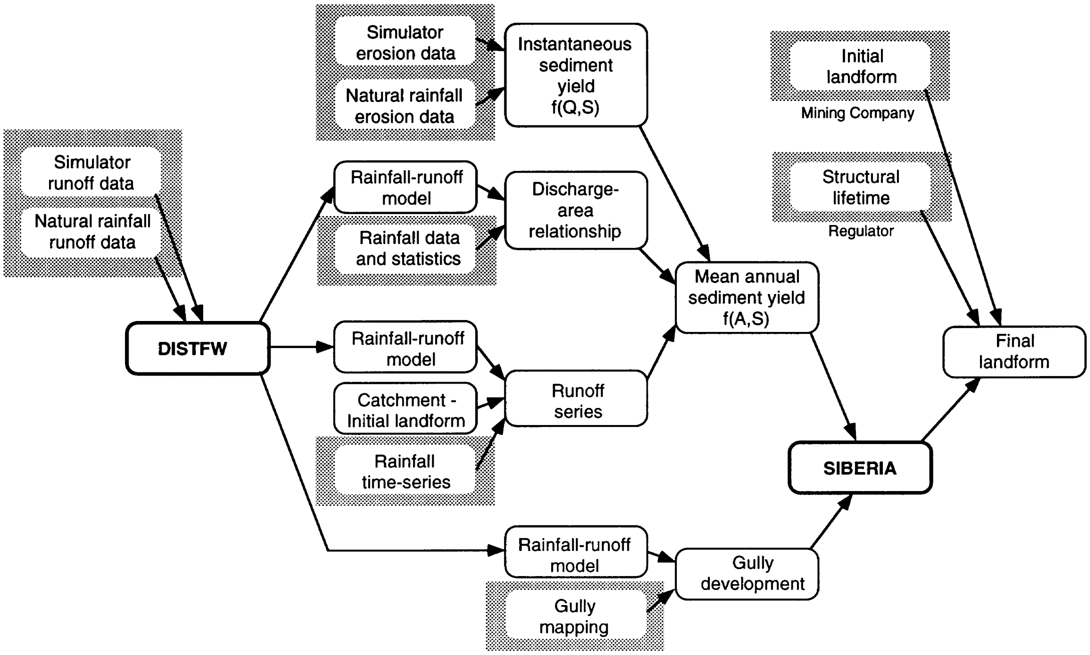


Figure 13. Schematic of the calibration process and use of SIBERIA

discharge with area for specified rainfall data. This model is then used to predict the scale dependence of the mean peak discharge: the variation of the discharge with increasing area and slope.

Finally, if the gully development module of SIBERIA is required, a gully threshold is required to predict the extent of potential gullying. Data for a nearby natural site with similar regolith properties were used to estimate the magnitude of this threshold and its dependence on hillslope gradient area. This calibration process is summarized in Figure 13.

Scale dependence of the hydrology

Some of the most important parameters in SIBERIA are those that define how the discharge used in the calculation of the sediment transport rate varies with catchment area. The relationship between discharge and area used in SIBERIA (Equation 3) has been widely used in empirical studies of catchment hydrology and is the basis of many regional relationships for flood frequency. Recently, Huang and Willgoose (1993) and Willgoose and Kuczera (1995) have studied how the DISTFW rainfall–runoff model may be used to determine this relationship for small catchments by applying spatially uniform rainfall and plotting peak discharges with area. This process is only valid for small catchments where it is reasonable to assume that the rainfall in all parts of the catchment is the same, but could be extended to larger catchments if spatial statistics of rainfall are available.

The process is as follows:

1. Calibrate or select the parameters for the DISTFW model.
2. From intensity–frequency–duration (IFD) curves of rainfall the two-year storms of various durations are selected. Using the rainfall temporal patterns from Australian Rainfall and Runoff (Institution of Engineers, Australia, 1987), each of these storms is applied uniformly over the catchments and the peak discharge for every node in the catchment is noted for each storm.
3. The peak discharge at each node from the various duration two-year storms is determined (smaller areas have highest discharge from short storms, larger from longer storms). These peak discharges are then plotted against area and the coefficients of the discharge–area relationship in SIBERIA are directly fitted from the graph. Huang and Willgoose (1993) found that the correlation coefficient of this relationship is very high,

and that the parameters in the relationship are a function primarily of the conveyance parameters in the rainfall–runoff model.

This process was followed to determine the area dependence of the discharge at RUM.

Long-term erosion rate and time-scales for the simulation

For the determination of the long-term erosion rate, a runoff series is created using the historical rainfall records at Jabiru and the calibrated rainfall–runoff model. Using the sediment transport equation previously calibrated and this runoff time-series, an erosion time-series is generated and the average sediment transport per year can be determined. Details are given in Willgoose and Riley (1993).

Gully thresholds

One of the novel features of SIBERIA is its ability to model the dynamic development of gullies in response to hydrologic and erosional characteristics of the surface. Using a user-defined threshold, a gully occurs when that threshold is exceeded by a function called the channel initiation function (CIF). This CIF is a function of the hydrology upstream of the gully head. This hydrology is, in turn, a function of upslope area and slope. The CIF can, for example, be a function of the velocity of the overland flow, the shear stress of the overland flow, or, in areas dominated by groundwater flow, a function of the groundwater head gradient. Most importantly, these functions are positively correlated with runoff and rainfall, area per unit width upstream of the gully head and the slope at the gully head (Willgoose *et al*., 1989). Everything else being equal, increases in rainfall, runoff, area and slope increase the tendency for a gully to be created at any given point in a catchment. This threshold behaviour based on area, slope and runoff has been widely observed in natural catchments (Patton and Schumm, 1975; Montgomery and Dietrich, 1989).

SIBERIA requires that this area–slope–runoff relation be determined *a priori* from field data and used as input parameter to the model. Once a gully is triggered, by exceeding the channel initiation threshold, the excavation of that channel proceeds using the sediment transport physics discussed earlier.

No data exist for the RUM rehabilitation site at this stage to allow the calibration of such a relationship. However, Riley and Williams (1991) have examined a natural area (Tin Camp Creek) derived from similar geologic material – schists.

The channel initiation function used by SIBERIA is:

$$\frac{a}{a_\mathrm{t}} = \frac{\beta_5 q^{m_5} S^{n_5}}{a_\mathrm{t}} \begin{cases} > 1 \text{ channel / gully} \\ < 1 \text{ hillslope} \end{cases} \tag{14}$$

This function is both consistent with field data (Willgoose *et al*., 1990) and justified from theoretical considerations (Willgoose *et al*., 1989; Dietrich, 1994). This relation can be re-expressed as:

$$\frac{a}{a_\mathrm{t}} = \frac{\beta_5' A^{m_5'} S^{n_5'}}{a_\mathrm{t}} \tag{15}$$

where the primed variables are functions of the parameters of the CIF and the discharge relationships. Discriminant analysis (Mosteller and Tukey, 1977) was used to identify the threshold between gullied and ungullied. The result is plotted in Figure 14. The resulting channel initiation function is:

$$\frac{a}{a_\mathrm{t}} = \frac{A^{2\cdot 27} S}{23\cdot 6 \times 10^6} \tag{16}$$

The power on area in the above expression is in the range of values that have been observed for natural catchments and within the range of values predicted from theory. For a slope of 0·15 (approximately that of the waste rock batters) it predicts an area of about 4000 m^2 which for a planar slope is a slope length of about 65 m. For a slope of 0·02 (approximately that of the waste rock cap) it predicts an area of about 10 000 m^2 which for a

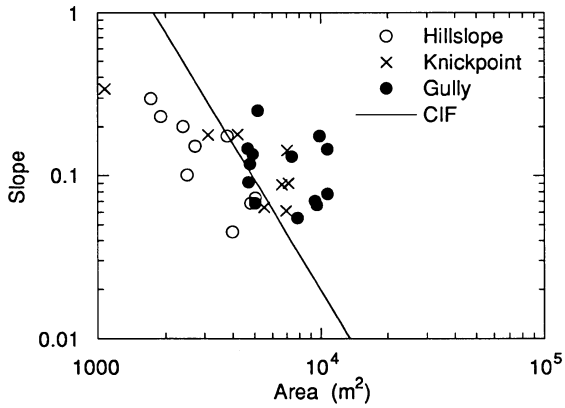


Figure 14. Adopted threshold for distinguishing gully from hillslope (data from Riley and Williams, 1992)

planar slope is a slope length of about 100 m. While Riley and Williams (1991) proposed a slightly different relationship it appears that the discriminating power of Equation 16 and theirs are similar.

ASSESSMENT OF PROPOSED RUM LANDFORMS

Above-grade option: without gullies

The above-grade option was run with SIBERIA for the equivalent of 1000 years using the parameters calibrated in the previous sections. Gullies are assumed not to develop in the bottom of valleys. Perspectives of the waste rock dump are for the as-constructed year zero condition and for the 500 and 1000 year cases (Figures 15 and 16). The grid used in this figure is a 60 m grid. Since the calculations were done on a 30 m grid only every second elevation value is plotted. This was done to simplify the visualization though inevitably some detail, particularly of the narrower valleys, is lost. While the differences in the surfaces for 0 and 1000 years may appear to be small, it must be remembered that the waste rock dump is almost 20 m high. The maximum valley depth is 7·7 m with the maximum deposition being 6·1 m. Note that while these valleys appear to be deeply incised in the figures, this is simply a function of the vertical exaggeration of the figures. Even for the maximum depth of 7 m over the 30 m grid resolution means that valley side slopes are still only 0·2, much less than for typical incised gullies.

Figure 15 also shows elevations at 500 years. Maximum erosion depth at 500 years is 5·7 m, 74 per cent of the maximum 1000-year erosion. This indicates that incision of the valleys proceeds most rapidly in early years.

Above-grade option: with gullies

The simulations reported above were repeated but allowing gullies to develop, as parameterized by Equation 16. In the absence of information of the sediment transport in the gullies at Tin Camp Creek, the sediment transport rate in the gullies was assumed to be equal to that on the hillslopes (e.g. SIBERIA's parameter $O_t = 1$). The sensitivity of the gully development to the random settlement of the waste rock (approximately 1 m on average) was also examined, particularly how the positions of gullies change with the imposition of settlements. We did not predict the depth of gully development (that would require further data from studies at Tin Camp Creek). However, the gully incision can be simply added to the depths of sheet erosion discussed above; the two depths of erosion are believed to be largely independent over geomorphic time-scales (Figure 17).

The gullies extend quite some distance into the central cap area, fanning out to fill the lower regions of the cap. They do not extend all the way to the drainage divide on the cap because there the contributing area and slopes are too low to trigger gully development. The gully development on the batters is largely constrained to the batters themselves although some extend onto the cap for a short distance in the southwest corner. That most gullies stop at the top of the batters is not surprising since the slopes abruptly decrease at this point and on the cap the CIF theshold is no longer exceeded. The extension onto the cap in the southwest corner of the batter is

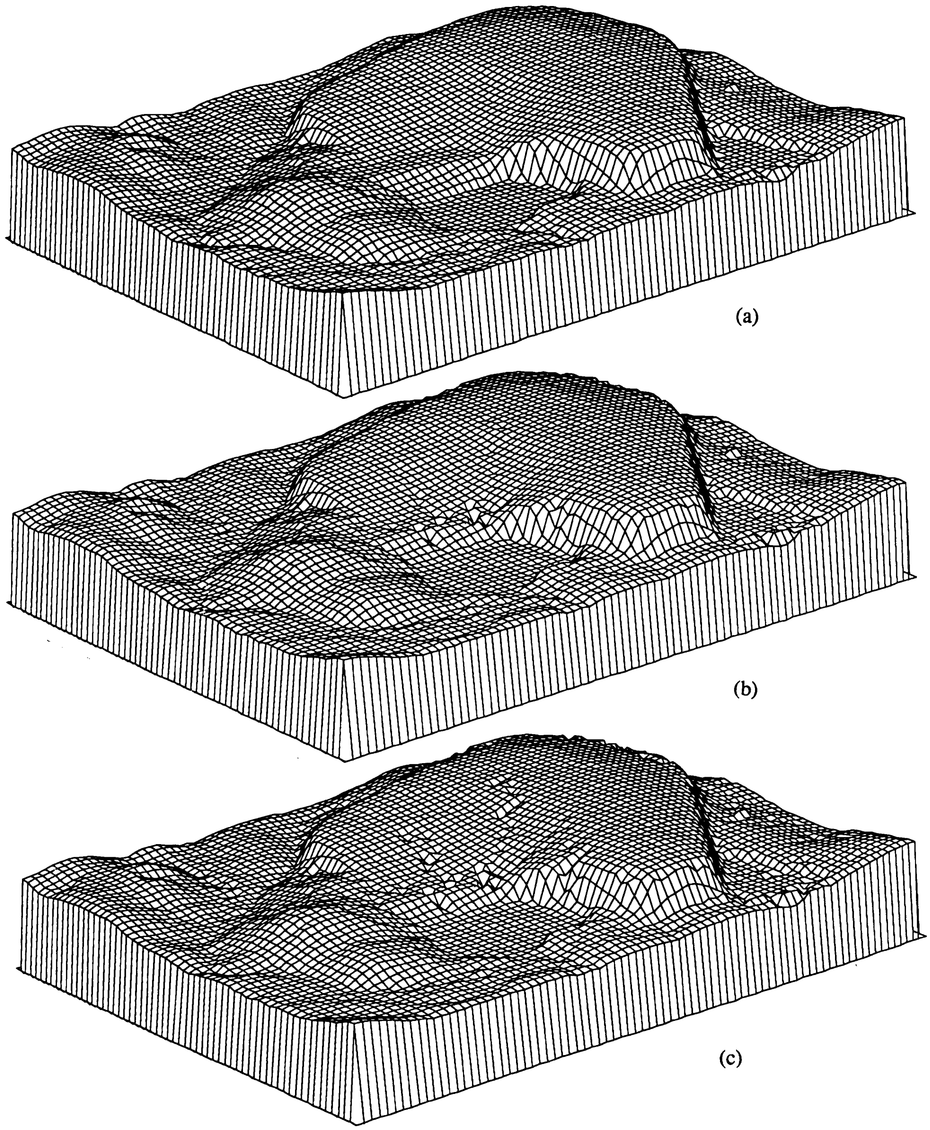
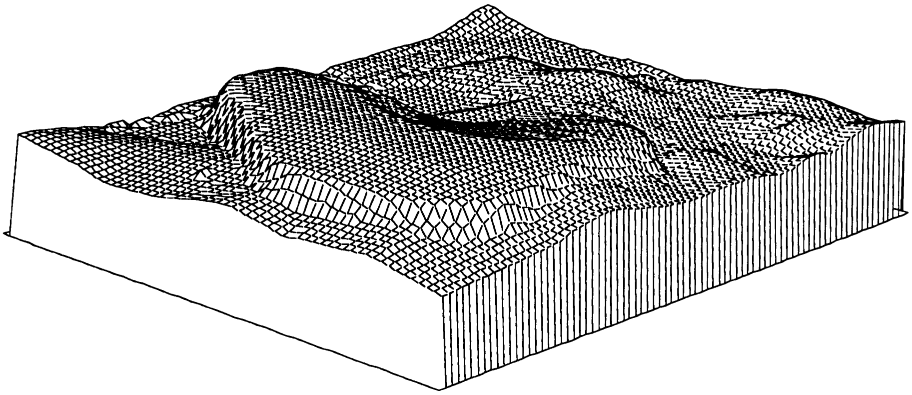


Figure 15. Above-grade option, baseline case: elevations at (a) 0, (b) 500, (c) 1000 years viewed from the NE

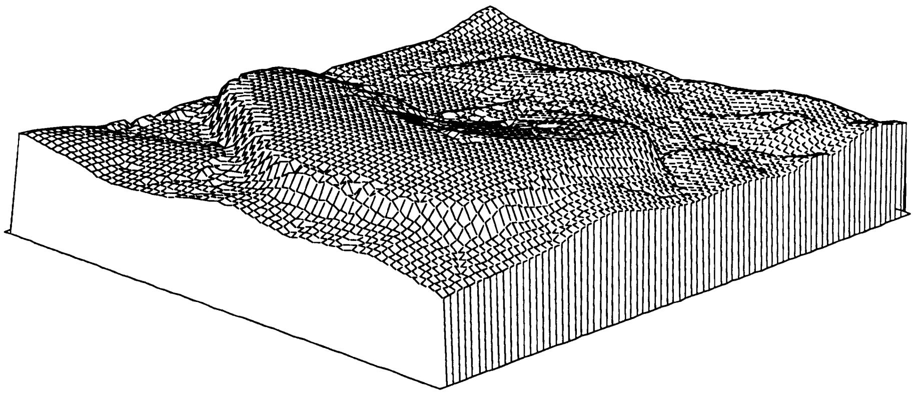
possible because of the longer slope lengths on the cap contributing to the batter at that point. However, even for the regions where gullies do not extend onto the cap in 1000 years, it is likely that over longer time-scales they will extend as valleys incise.

LONG-TERM VERSUS SHORT-TERM EROSION MODELLING

Traditional methods of erosion assessment, such as USLE, RUSLE and CREAMS, determine the erosion at any given time for a particular landform. They are not able to predict the change in shape of the landscape as a result of the erosion that occurs on it, nor are they able to predict the effect that the change in the landform has on the future erosion patterns. These methods are implicitly short-term techniques. If erosion predictions are only



(a)



(b)

Figure 16. Above-grade option, baseline case: elevations at (a) 0, (b) 1000 years viewed from the SW

required for a small time in the future, over which the erosion does not change the landform much, then they provide good predictions of erosion patterns.

Over longer periods of time, however, the change of landform shape cannot be ignored. This is the rationale for the use of the SIBERIA landscape evolution model for the erosion assessment in this report. Localized erosion results in localized convergence of flow with further increases of erosion. Thus valleys will deepen and widen over time as the natural process of drainage development occurs. While over the short-term the predictions of the short-term and long-term models will be little different, as the landform erodes the short-term models will progressively provide poorer estimates of the erosion. In particular, the spatial pattern of erosion, where the localized high erosion occurs on the rehabilitation, will be poorly estimated.

Figure 18 demonstrates the difference between the two modelling approaches, showing the patterns of erosion for three cases for the above-grade baseline proposal. The first case, Figure 18a, shows the pattern of erosion for the first three years after the rehabilitation is complete (assuming that the surface of the rehabilitation was completed at the same time, rather than progressively during mine operation); the second case, Figure 18b, shows the pattern of erosion for the three years after the end of the design life (i.e. year 1000 to 1003). Both of these cases can be considered to be indicative of the results from short-term modelling exercises

(a) (b)

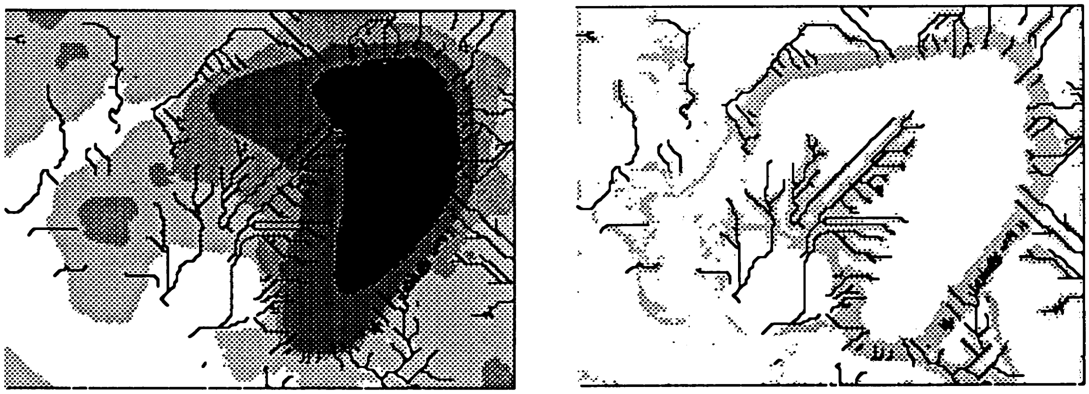


Figure 17. Above-grade option, baseline case with channels: channel predictions at 1000 years overlaid on grey-scale contours of (a) elevation and (b) slope

(a) (b) (c)

Figure 18. Comparison of the patterns of erosion for the short-term and long-term erosion modelling methods (see text for discussion)

with specified landforms. The second case, of course, assumes that we know the result of the long-term modelling exercise. The third case, Figure 18c, shows the pattern of erosion for the 1000-year long-term modelling. The pattern of erosion is different in both cases. Both of the short-term results predict a more uniform erosion distribution than the well defined, localized erosion apparent in the long-term result.

The 1000-year short-term result exhibits some localized erosion, not only in the valleys already created by the long-term model; it does not predict the localized erosion upstream that will probably occur after 1000 years in the long-term case. The reason lies in the way valleys are incised over time. Valleys do not gradually downcut over their whole length with time. If they did, then erosion depth estimates from the short-term models could be

factored up for the design life of the landform. Rather, valleys rapidly incise at the valley head as it propagates upstream from the highwall around the central retention pond with proportionally less erosion both upstream and downstream of the headcut. This is apparent in Figure 18c where the highest erosion rates are at the valley heads. Short-term methods can predict where the areas of localized high erosion will occur at any given time for any specified landform; however, they are unable to predict how this region of localized high erosion will move over time. They are thus unable to accurately estimate the spatial distribution of the regions of high erosion.

While the pattern of erosion is different in the short-term and long-term results, the average rate of erosion over the domain is much the same (about 0·3 m over the 1000 years). The difference between the short-term and long-term results is that much more of the long-term erosion is concentrated in deep valleys; the short-term results would predict more uniform, less concentrated erosion depths. In short, the short-term modelling results would be non-conservative with lower values for maximum erosion depths, even though average depths of erosion appear to be similar for both short-term and long-term results.

CONCLUSIONS

The simulations herein have clearly shown that significant erosion will occur in the next 1000 years in the cap region of both the above-ground and below-ground options. Peak erosion depths without gully development are predicted to be in the range of 7–8 m. Gully development potentially increases the maximum penetration of the cap layer further. It is predicted that a number of valleys will dissect the central region of the cap rock. The exact position of these valleys is subject to some doubt because of the poor definition of an initial drainage structure on the proposed designs. It thus appears difficult to design localized protective measures for these gullies because the position of these potential gullies cannot be predicted *a priori*. Drainage network development is a chaotic process (Willgoose *et al*., 1991c; Ijjasz-Vasquez *et al*., 1990) but if an initial drainage pattern is imposed, some predictability should be imposed on the eroding system.

In addition, it has also been shown that the steep (slope = 0·15) batter slopes will suffer severe degradation of the order of 5–7 m. The valleys on the batters do not occur in predictable places but occur along all the batter extremities of the waste rock dump. The erosion problem is thus not localized to one place, where it potentially may be protected, but it occurs across broad areas making it difficult to design reliable protective measures. The fundamental cause of this problem is that there are substantial slope lengths (>200 m) on the cap that contribute flow to the upper end of the batters. When this flow reaches the batter it cascades over the batter causing severe degradation. One solution – bund walls around the top of the batter – is unlikely to solve this problem. Moreover, because of the risk of localized overtopping and consequent concentration of flow, the widespread nature of the erosion on the batters indicates that there appear to be few safe locations to which this flow can be diverted.

The substantial erosion on the batters results in deposition in the surrounding areas. The deepest depths of deposition (about 5 m over 1000 years) appear to very close (within 150–200 m of the batters) to the batters, although it is apparent that some deposition does occur at greater distances. More precise location of deposition area would require erosion parameters for the undisturbed areas surrounding the engineered structures.

Finally, in the absence of random settlement, the rates of erosional loss on the majority of the cap away from the gullies appears to be relatively small (less than 500 mm). In fact, the low erosional loss on the portion of that cap contributing to the batters enhances the gully erosion that occurs on the batters. Addition of random settlement with a range of 0–1 m induces erosion and deposition on the cap of about 1 m depth. There is no apparent systematic pattern to this erosion.

This deep sheet erosion in isolated regions, with little erosion in intervening areas, suggests a solution strategy for the problem and involves considering the geomorphology of the entire waste rock dump. As previously noted, the major problem with the existing designs is that their slopes decrease as drainage area increases in different fashion to that observed in natural catchments, which are closer to their equilibrium form. This characteristic of natural catchments arises from the balance of erosion, drainage patterns and elevations that catchments tend towards over geological time-scales (Gilbert, 1909; Willgoose *et al*., 1991d; Willgoose, 1994).

Thus, in summary, there are three problems with the proposed rehabilitation structures which should be addressed.

The first problem is that the slope gradient does not decrease downslope as it does in natural catchments. This feature means that sediment transport increases much faster downslope than occurs in natural catchments. The long-term effect is for gully erosion to develop at the bottom of the slopes as the lower parts of the catchments trend towards the low slope condition.

The second problem is that the wide flat hillslopes allow gullies to concentrate flow (and thus increase the discharges and erosion) with great ease. By imposing a drainage network it becomes very difficult for a gully to capture adjacent areas (first they must erode away the interfluve). Discharges are then unlikely to change as erosion proceeds. A secondary advantage is that if a gully does occur it will be localized and its growth will be controlled. A key feature controlling the rate of growth of gullies is their ability to capture area; reduce this ability and gullies grow less quickly.

The third problem is that the long cap hillslopes contribute flow to the tops of the batters, inducing deep erosion at the tops of the batters.

ACKNOWLEDGEMENTS

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